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I. Summary of Accomplishments

This research project involved the investigation the vertical profiles of temperature and moisture in convective regimes, using moist available energy as a guide. The results have been used to develop an improved cumulus parameterization.

From a human perspective, kinetic energy is the most important energy form of the atmosphere. All weather systems owe their existence directly to the kinetic energy that they possess. Whether a weather system is intensifying or weakening depends on if it is gaining or losing kinetic energy. Therefore, the source or sink of kinetic energy is a matter of importance. When the source of kinetic energy is huge and increasing, we can predict that the associated weather system will develop and persist and what intensity it will reach; on the other hand, when the source of kinetic energy is limited and decreasing, we can conclude that the associated weather system will weaken. Under adiabatic and frictionless conditions the total energy of the atmosphere, which is the sum of its kinetic energy, potential energy and internal energy, would remain constant. In such a case, the only sources or sinks for the kinetic energy of the whole atmosphere would then be potential energy and internal energy.

Generally speaking, however, the motion of the atmosphere is neither adiabatic nor frictionless. The most important nonadiabatic process which directly alters the atmospheric kinetic energy is friction, which ordinarily generates internal energy or does work on the ocean to increase the ocean's kinetic energy, while destroying the kinetic energy of the atmosphere. Since the intensification of a weather system is often directly related to the production of its kinetic energy, the loss of kinetic energy due to friction is not so important compared to the kinetic energy production. This is especially true when we deal with a short time scale, for instance, several hours or days. As a result, we can sometimes usefully consider a frictionless atmosphere. The other nonadiabatic processes do not alter the kinetic energy of the atmosphere directly, but only alter the internal energy. Therefore, the effects of these processes on the atmosphere can be reflected through the change of internal energy. Hence, after assuming frictionless motion, the only sources for the kinetic energy of the whole atmosphere are its potential energy and internal energy.

For a column of air standing from the surface to the top of the atmosphere, the vertically integrated total potential energy, defined as the sum of potential energy and internal energy, equals the vertically integrated enthalpy. For the whole atmosphere, the sum of the enthalpy and kinetic energy is conserved under adiabatic and frictionless conditions. This means that the only source of kinetic energy is the enthalpy of the atmosphere. When the enthalpy of the atmosphere decreases adiabatically, the kinetic energy increases.

As discussed by Lorenz (1955), however, the enthalpy (or the total potential energy) is not a good measure of the amount of energy available for conversion into kinetic energy under adiabatic and frictionless flow, since in the atmosphere, not all of the enthalpy can be converted into kinetic energy. A simple example was given by Lorenz (1955) to illustrate this point. Consider first an atmosphere whose density stratification is everywhere horizontal. In this case, although total enthalpy is plentiful, none at all is available for conversion into kinetic energy. Next, suppose that the horizontally stratified atmosphere becomes heated in a restricted region. This heating adds the total enthalpy of the atmosphere, and also disturbs the stratification, thus creating horizontal pressure forces which may convert enthalpy into kinetic energy. On the other hand, suppose that the horizontally stratified atmosphere becomes cooled rather than heated. The cooling removes enthalpy from the system, but it still disturbs the stratification, thus again creating horizontal pressure forces which may convert enthalpy into kinetic energy. Evidently cooling is sometimes as effective as warming in producing kinetic energy, and the total enthalpy itself is not a good measure of how much enthalpy is available for conversion into kinetic energy. It seems that only that portion of enthalpy which can be increased or decreased by the atmospheric motion can be used as the source of kinetic energy.

We therefore desire a quantity which only measures this portion of total potential energy which is available for conversion into kinetic energy under adiabatic and frictionless flow.

According to Lorenz (1955), a quantity of this sort was first discussed by Margules (1903) in his famous paper concerning the energy of storms. Margules considered a closed system possessing a certain distribution of mass. Under adiabatic flow, the mass may be redistributed, with an accompanying change in total potential energy, and an equal and opposite change in kinetic energy. If the stratification becomes horizontal and statically stable, the total potential energy reaches its minimum possible value, and the kinetic energy thus reaches its maximum. This maximum gain of kinetic energy equals the maximum amount of total potential energy available for conversion into kinetic energy under any adiabatic redistribution of mass, and therefore was called "available kinetic energy" by Margules.

The concept of "available potential energy," which is similar to the "available kinetic energy" defined by Margules, was introduced by Lorenz (1955) in considering the general circulation of the atmosphere. The available potential energy (APE) of the whole atmosphere is defined as the difference between the total potential energy of the whole atmosphere and the minimum total potential energy that the whole atmosphere would have if the mass were redistributed adiabatically to yield a horizontally uniform and vertically stable stratification.

As demonstrated by Lorenz (1955), the APE, so defined, possesses the following important properties:

(1) Under adiabatic flow, the sum of the available potential energy and the kinetic energy is conserved. The APE is the only source of kinetic energy, but it is not the only sink.

(2) The available potential energy is completely determined by the distribution of mass.

(3) The available potential energy is zero if the stratification is horizontally and statically stable. Also, the APE is positive if the stratification is not both horizontally and statically stable.

The "reference state" is defined as the state in which the atmosphere has the minimum total enthalpy that could be reached by rearranging the mass under reversible adiabatic processes.

Then, for any given state of the atmosphere, the APE is defined as the enthalpy difference between the given state and its reference state.

Lorenz (1955) discussed in detail the APE of the whole atmosphere. Although solar radiation is the ultimate source of the atmospheric energy, the atmosphere does not obtain its energy only through solar radiation. It also obtains energy from terrestrial radiation, latent and sensible heat fluxes from the Earth's surface. The globally and annually averaged energy balance (e.g., Peixoto and Oort, 1992; the specific numbers below are from Ramanathan, 1987) shows that the largest energy source for the atmosphere is the latent heat flux from the surface, mainly the ocean surface, with 90 W m^{-2}. The atmospheric absorptions of solar and solid earth radiation, 68 and 63 W m^{-2} respectively, are smaller, compared to the latent heat flux from the ocean surface; while the surface sensible heat flux is the smallest atmospheric energy source, with 16 W m^{-2}.

An essential feature of the solar radiation as received by the Earth is that it is horizontally non-uniform. And of course, the Earth's surface is also non-uniform. Because of these non-uniformities, the latent and sensible heat fluxes from the surface into the atmosphere and the heating and cooling of the atmosphere due to solar and terrestrial radiation are larger in the tropics than in the higher-latitude regions. As a result, a temperature contrast between the equator and the poles is produced. The unbalanced pressure forces demanded by the temperature contrast produce a circulation to transport energy from the region of net energy gain to the region of net energy loss. Thus, most of the APE in the atmosphere is associated with the horizontal temperature contrast which is generated by the horizontally non-uniform energy supply.

As discussed above, the APE is that portion of the total potential energy which can be converted into kinetic energy. There is no assurance that all of the APE will be converted into kinetic energy, however. How much of the APE will be converted varies from case to case. For the whole atmosphere, Lorenz (1955) estimated that the amount of kinetic energy is only about 10% of the amount of APE. Evidently, if kinetic energy is not fully maximized, it is not because a supply of APE is lacking, but because there are not dynamically realizable circulations that can extract all of the APE.

In Lorenz's (1955) study, however, the role of moisture in the APE was not discussed specifically. Since most of the Earth's surface is covered by the oceans, the evaporation of sea water leads to a large amount of water (mostly in the vapor state) in the atmosphere. Many of the more spectacular weather events, from tropical hurricanes to polar blizzards, owe their existence to the latent energy of condensation and the fusion of water in the atmosphere. Therefore, it is more reasonable to deal with a moist atmosphere than a dry atmosphere in consideration of the APE.

The greatest difficulty in dealing with the APE of moist atmosphere is to include the effects of the latent heat of condensation and fusion of water vapor in the definition of the APE. In the real atmosphere, when condensation happens, latent heat is released, and some of the condensate will drop out as precipitation. This precipitation process is, of course, nonadiabatic, whereas the APE is defined in terms of adiabatic processes. If we assume, however, that all the condensate accompanies the air in which it condenses, the process is still adiabatic. In such a case, the sum of enthalpy and kinetic energy is still conserved, and therefore, the concept of APE is still applicable. Based on this idea, Lorenz (1978,1979) extended the concept of APE to the moist atmosphere. With the effects of water condensation included in the definition of the enthalpy, the APE was called the moist available energy (MAE) by Lorenz (1978,1979), in contrast with the "dry available energy" (DAE) which does not include moisture effects. Lorenz (1978,1979) showed, through both graphical and numerical methods, that the MAE is always larger than the DAE, due to the condensation of some water vapor and the much smaller enthalpy of condensed water compared to water vapor. This shows that the latent heat of water vapor represents an additional source of atmospheric kinetic energy.

When the mass of the whole atmosphere is rearranged adiabatically from the given state to the reference state, the rearrangement is both horizontal and vertical. The horizontal rearrangement is needed to eliminate the horizontal pressure differences and temperature contrasts, while the vertical rearrangement is needed to maximize the static stability. The horizontal rearrangement of mass drives planetary and synoptic circulations whose time scale is days; whereas the vertical rearrangement mostly drives convection whose time scale is several hours. Therefore, the "horizontal part" of the MAE is not effectively accessible to cumulus convection. The "vertical part" of the MAE, however, can be a source of kinetic energy for cumulus convection. Thus, the concept of MAE can be applied to a column of air to measure the convective instability that it possesses. For such an air column, the "vertical component" of the MAE is a measure of the portion of total potential energy available for conversion into convective kinetic energy, and is similar to the Convective Available Potential Energy (CAPE). In next chapter, we will define such a quantity for an atmospheric column, as a measure of its CAPE, and compare the atmospheric instability as measured by this new method with those from other methods.

When we lift an air parcel to an arbitrary new level, if its density is greater than that of the surrounding air, it will tend to return to its original position. Such an atmosphere is said to be statically stable. Otherwise, if the density of the lifted parcel is less than that of the environmental air at the new level, the lifted parcel will be positively buoyant and will, therefore, tend to accelerate upward. In such a case, the atmosphere is statically unstable. The neutral state is the "boundary" between the stable and unstable states. In the neutral state, the density of lifted parcel is equal to that of the environmental air.

Since the approximation that a lifted air parcel has the same pressure as its surrounding environment can be used, we also can use the temperatures of a lifted parcel and environment to judge the static stability of the atmosphere. If the lifted parcel is warmer (colder) than its environment, it is lighter (heavier) than the environment, and therefore the atmosphere is statically unstable (stable). Strictly speaking, when we consider a moist atmosphere, virtual temperature should be used instead of temperature to measure the static stability.

Since the lapse rate represents the rate of temperature change with height, it can be used to measure the static stability of the atmosphere. For a dry atmosphere, if the lapse rate ($\Gamma = -\frac{dT}{dz}$) is larger (cooling more rapidly upward) than the dry adiabatic lapse rate (Γ_d), that is, if $\Gamma > \Gamma_d$, then an air parcel lifted dry-adiabatically will become warmer than its environment so that the atmosphere is statically unstable. Similarly, if $\Gamma < \Gamma_d$, the atmosphere is statically stable.

For a given pressure, air with a higher (lower) temperature has a higher (lower) potential temperature (θ). And also, for dry adiabatic motions, the θ of a lifted parcel is conserved. This means that, for a dry atmosphere, by comparing the θ's of two parcels, we can determine which one will be warmer or colder at any given pressure level. The air with the higher θ will be warmer than the air with lower θ, when they are put at the same pressure. Therefore, the vertical derivative of θ can be used as a measure of the static stability. When $\frac{d\theta}{dz} > 0$, the θ of lower-level air is less than that of higher-level air, so when the lower-level air is lifted, it will be cooler than its environment. The atmosphere is therefore statically stable. Similarly, when $\frac{d\theta}{dz} < 0$, the atmosphere is statically unstable, and when $\frac{d\theta}{dz} = 0$, the atmosphere is statically neutral.

In reality, however, the atmosphere contains water vapor. When a parcel is lifted, it may become saturated, so that some water vapor condenses, and latent heat is released. The latent heating raises the temperature of the lifted parcel. The static stability criterion for a moist atmosphere is, therefore, different from that of a dry atmosphere.

One of the simplest concepts used in the analysis of a moist atmosphere is the "pseudoadiabatic" process, in which all condensed water is assumed to drop out as soon as it forms. The pseudoadiabatic lapse rate (Γ_s) can be used to judge the static stability of the moist atmosphere. If the atmospheric lapse rate (Γ) is larger (cooling more rapidly upward) than the pseudoadiabatic lapse rate (Γ_s), that is, $\Gamma > \Gamma_s$, the atmosphere is statically unstable for pseudoadiabatic motions. Otherwise, if $\Gamma < \Gamma_s$, the atmosphere is statically stable for pseudoadiabatic motions.

It can easily be shown (e.g. Holton, 1979) that the dry-adiabatic lapse rate is always larger than the pseudoadiabatic lapse rate, that is, $\Gamma_d > \Gamma_s$, because of the latent heat of condensation. When an atmosphere is statically stable for dry adiabatic motions, it may statically unstable for pseudoadiabatic motions. This is just the case in which the lapse rate lies between Γ_s and Γ_d,

$\Gamma_s < \Gamma < \Gamma_d$, so that the atmosphere is stable with respect to dry adiabatic displacements but unstable with respect to pseudoadiabatic displacements. Such an instability is referred to as "conditional instability." The "condition" is the saturation of the lifted parcel.

In the real atmosphere, the lapse rate is rarely greater than the dry adiabatic lapse rate, so that the atmosphere is mostly statically stable for dry adiabatic processes. The most common type of static instability is the conditional instability. It can be shown (e.g. Holton, 1979) that θ_e, the equivalent potential temperature, can be used approximately to define a criterion for the conditional stability. When $\frac{\partial\theta_e}{\partial z} < 0$, the atmosphere is unstable; $\frac{\partial\theta_e}{\partial z} > 0$, stable; $\frac{\partial\theta_e}{\partial z} = 0$, neutral.

If the atmosphere is statically unstable, a lifted moist parcel will obtain positive buoyancy and therefore convective kinetic energy. It is therefore convenient to measure the atmospheric instability by the convective kinetic energy that the parcel will obtain. The Convective Available Potential Energy (CAPE) is just such a measure. It represents the maximum possible kinetic energy that a lifted parcel can acquire in a conditionally unstable atmosphere. The kinetic energy that a lifted parcel can acquire is the work done by the buoyancy force, i. e.

$$CAPE = \int_{p_{NB}}^{p_{FC}} R_d (T_{v,p} - T_v)\, d\ln p. \tag{1.1}$$

Here p_{FC} is the pressure of the free convection level, p_{NB} is the pressure of the neutral buoyancy level, R_d is the specific gas constant of dry air, $T_{v,p}$ is the virtual temperature of the lifted parcel, and T_v is the virtual temperature of the environment. CAPE is an energy-related measurement of convective instability. The cloud work function of Arakawa and Schubert (1974) is a similar energy-related concept. It also measures the kinetic energy that a lifted parcel can obtain through the work done by buoyancy. Unlike the CAPE, however, the cloud work function includes the effects of entrainment on buoyancy.

The above are the traditional methods to measure the static instability of the atmosphere or the convective kinetic energy that a lifted parcel can acquire in a statically unstable atmosphere. One of the basic assumptions of these methods is that the lifted parcel is small enough so that its displacement does not disturb the surrounding environment. In reality, however, the rising motion must be compensated for by subsidence in the environment, if an overall mass balance is to be maintained. The dry environment between the cloudy updrafts must descend, so that the environment is in fact disturbed. Thus, we should try to take into account the changes of the environment when the instability of the atmosphere is considered.

Bjerknes (1938) first included the effects of the environmental air in the computation of conditional instability. He calculated the conditional instability of the adiabatic ascent of saturated air through a dry-adiabatically descending environment. He assumed that on any horizontal plane the upward mass flux in convection cells is just balanced by the downward mass flux in the cloudless environment. The saturated air ascends adiabatically, while the environment descends dry-adiabatically. He showed that the net heating for a layer can be expressed as the sum of two terms, one representing the released latent heat of condensation in clouds, and the other representing the dry-adiabatic warming due to the downward compensating flow. The latent heat of condensation in clouds was considered to be the only source of convective kinetic energy.

Because of the warming of the environment due to the compensating dry-adiabatic descent, a rising parcel must be heated more than if the environment were undisturbed, in order to obtain a given amount of buoyancy. Bjerknes' results showed that in order for the rising cloudy parcel to obtain positive buoyancy, the cloudy tower must be narrow enough. That is, convective clouds are likely to occur in a system with appreciable upward velocity in narrow cloud towers and slow downward motion in the wide cloudless spaces. How narrow the cloudy area must be, compared to the cloudless area, depends on the lapse rates of the observed sounding and the dry- and saturated-adiabatic processes.

Bjerknes also compared two measures of convective instability, i.e., the classical method in which an infinitely small parcel does not disturb its environment, and his method in which finite cloud towers lead to environmental sinking. For this purpose, he used an ordinary sounding. For such a sounding, when a parcel is lifted saturated-adiabatically from its saturation level to a point above, the parcel method shows that the lifted parcel will have positive buoyancy and gain convective kinetic energy; but Bjerknes method shows that the finite lifted parcel will have negative buoyancy and cannot obtain convective kinetic energy at all. Bjerknes also did other comparisons between the two methods. His conclusion is that the atmosphere is always less unstable with respect to a system of finite cloud towers than with respect to the infinitely small saturated parcels.

As has been shown by Bjerknes, the effects of compensating return flow between clouds can change our conclusion about the degree of convective instability. Although Bjerknes showed the importance of including the compensating return flow, he did not quantitatively show how to measure the convective instability with the return flow included. We will show in next chapter that the method that we propose does include the effects of compensating return flow quantitatively for measuring convective instability of the atmosphere.

In addition, the previous parcel-lifting methods for measuring the convective kinetic energy (e.g., CAPE, cloud work function) depend on the choice of the parcel lifted. For different lifted parcels, the measured instability may be different. This can be clearly seen, for instance, from the definition of CAPE, (1.1). It shows that the CAPE is the total work done by buoyancy when a parcel is lifted. The value of this work is the product of two factors: the buoyancy that the lifted parcel experienced, and the length of the path over which the parcel has positive buoyancy. If a warmer and wetter parcel is lifted, it will have more buoyancy and its path will be longer, so its CAPE will be larger than that of a colder and / or drier parcel. This shows that for a given sounding, lifting different parcels can give different CAPEs. It is clearly desirable to have an unique value of CAPE for a given sounding. We will show that our measure of the convective instability is unique. It is a property of a whole atmospheric column, and does not make reference to any particular lifted parcel.

Cumulus convection, especially deep and intense convection, is one of the major processes affecting the dynamics and energetics of large-scale atmospheric circulations. The ways through which convection exerts influence include: diabatic heating due to latent heat release in penetrative cumulus convection; vertical transports of heat, moisture and momentum; and the interaction of cumulus clouds with radiation. Riehl and Malkus (1958) showed the importance of cumulus convection for the heat balance of the tropical atmosphere. They showed that deep cumulus convection carries the released latent heat of condensation to the upper troposphere, to balance radiative cooling there.

The role of deep convection in the formation and growth of tropical cyclones was discussed by Riehl and Malkus (1961), Yanai (1961a,b), Ooyama (1964), and Charney and Eliassen (1964). Their results showed that tropical cyclones largely owe their existence to the release of latent heat in cumulus convection. They also showed that, to appropriately explain the growth of tropical cyclones in any model, cumulus heating must be adequately parameterized in the framework of the large-scale motion.

Cumulus convection is thus very important for the large-scale flow. It is necessary to account for these effects in a quantitative way in models of large-scale circulations. Ideally, if the resolution of models were sufficiently fine, individual clouds and their effects on the environment could be calculated directly, so that we would not need to parameterize them. To do so, however, horizontal and vertical grid sizes of between 100 and 1000 m would be required (Cotton and Anthes, 1989). Such a high resolution covering the domain size necessary to simulate larger-scale phenomena is far beyond present and foreseeable computational capability. The problem of cumulus parameterization will, therefore, remain important in any foreseeable future. Even with sufficient computer power, cumulus parameterization is still useful for understanding. Moreover, demands for simple numerical or theoretical models always exist regardless of computer power.

Before parameterizing the effects of cumulus convection, we need a throough understanding of the structure and dynamics of individual clouds and the micro behavior of these clouds. Unfortunately, our knowledge of clouds is quite limited, because the transports of mass, moisture, heat and momentum by clouds are not directly measured. Instead, we only can estimate what these transports must be in order to account for the residuals in the large-scale budget equations.

To do so, a diagnostic cumulus cloud ensemble model must be employed and therefore, the results are model-dependent.

Yanai *et al.* (1973) performed the first such study to determine the bulk properties of tropical cloud clusters from the large-scale heat and moisture budgets, using the Marshall Islands data. A bulk model was used. In this model, clouds were classified according to the heights of their tops, and all clouds were assumed to have the same cloud base. A system of equations, based on consideration of the cloud properties and the cloud effects on the large-scale fields, was solved to obtain the averaged cloud properties. The vertical profiles of these quantities describe the mean structure of the cumulus ensemble and the net effects of the clouds on their environment. The results showed that the upward mass flux in active cumulus clouds is larger than that required from large-scale horizontal convergence, thus causing a compensating sinking motion between active clouds. Entrainment, which is the mass added into the cloud from the environment from the side and / or the top of the cloud, was shown to be strongest in the lower troposphere, while detrainment, which is the mass carried away from the cloud into the environment from the sides and / or the top, has strong maxima in both the lower and upper troposphere. The lower detrainment maximum suggests the existence of a large number of shallow cumulus clouds in the region, co-existing with deep cumulus clouds. The large-scale heating of the environment by cumulus clouds was found to be primarily due to the adiabatic compression due to compensating downward motion. The cooling due to re-evaporation of liquid water detrained from clouds is also an important factor in the heat balance of the environment. The environment was found to be mostly dried by deep convection, due to the downward compensating flow between clouds. Counteracting the drying due to the environmental sinking motion are the large amount of water vapor and liquid water which are detrained from clouds, especially from the shallow clouds in the lower troposphere.

Also using the Marshall Islands data, Ogura and Cho (1973) applied a model of a cumulus ensemble to determine the cloud properties and the cloud contributions to the changes of heat and moisture content of the large-scale environment, from the large-scale budgets. Unlike the cumulus ensemble model of Yanai *et al.* (1973), which only estimated the average properties of the clouds, the Ogura-Cho model employed a spectral cloud model so that the properties of different cloud types could be found. This is one important advantage of the spectral model over the bulk model. The key point in the spectral model is that the properties of a single cloud are uniquely determined by an entrainment parameter, so that, by distinguishing the entrainment parameters, different cloud types can be distinguished. Arakawa and Schubert (1974) discussed the same spectral cloud model. The results of Ogura and Cho (1973) are basically the same as those of Yanai *et al.* (1973), except that the contributions from different cloud types can be seen.

As we have mentioned before, some observational studies showed the existence of downdrafts in cumulus clouds as early as the 1940's and 1950's (e.g., Byers and Hull, 1949; Squires, 1958). When we use cloud models to determine the properties of clouds, the effects of downdrafts should be included. Johnson (1976) incorporated downdrafts into the spectral cloud model. He assumed that each individual cloud element possesses an updraft and downdraft that are steady, entraining plumes, and that a constant ratio exists between the intensities of the updraft and downdraft. His results showed that the neglect of cumulus downdrafts and their associated rainfall evaporation leads to excessively large populations of shallow cumulus clouds in the highly convective situations. Nitta (1977) obtained the same conclusions by using an improved treatment of downdrafts in the spectral cloud model.

Since cumulus convection has important effects on larger-scale motions, and the grids of the large-scale models can not resolve the cumulus activity, we must find a way to relate the effects of the "subgrid-scale" cumulus clouds to the motions resolvable to the models. This is known as cumulus parameterization. For cumulus convection to be parameterizable, it is necessary that the convection be controlled by the large-scale motions. Many observations show that the large-scale processes do control convection.

According to Cotton and Anthes (1989), one of the first observational studies that showed a strong dependence of deep cumulus convection on larger-scale variables in the tropics was by Malkus and Williams (1963). They found that deep cumulus convection occurs only where low-level synoptic scale convergence prevails. They also noted that dynamic, rather than thermodynamic, factors are more crucial for cloud growth in the tropics, and that deep convection is characterized by marginal instability and low-level convergence, while very fair conditions are characterized by much stronger instability and divergence. These early observations were later confirmed by other studies (Matsumoto *et al.*, 1967; Cho and Ogura, 1974).

In addition, Cho and Ogura (1974) found a high correlation between the vertical mass flux in the deep clouds and the large-scale mass flux at 950 mb, for the composite wave data in the equatorial western pacific. Yanai *et al.* (1976) showed that deep cumulus clouds in the Marshall Islands region were highly correlated with the large-scale vertical motion in the upper troposphere. Nitta (1978) also showed that deep cumulus convection over the Global Atmosphere Research Program's Atlantic Tropical Experiment (GATE) area was highly correlated with the large-scale vertical velocity at all levels.

For the extratropics, there is also a lot of observational evidence showing that cumulus convection is strongly controlled by large-scale processes. For example, Sasaki and Lewis (1970), Lewis (1971), and Hudson (1971) found a close agreement between active convection and areas of mass and moisture convergence over the central United States. All of this observational evidence that deep convection is influenced and controlled by the large-scale motions provides a physical basis for cumulus parameterization, and implies that it is possible to parameterize the effects of cumulus convection in both the tropics and extratropics.

One of the purposes of cumulus parameterization is to determine the total net rate of condensation, and to distribute the convective heating and moistening vertically. Since the convective processes are subgrid scale, all cumulus parameterizations require closure assumptions. Since observations show that deep convection is invariably related to upward vertical motions and low level convergence, some authors have assumed that the large-scale vertical motion at the top of boundary layer and the low level convergence of mass and water vapor are proportional to some measure of the convective activity. Also, cumulus convection acts to release the convective instability in the atmosphere, so as to modify a conditionally unstable atmosphere toward a more stable state. If the approximate end-state can be specified, this can provide a convenient basis for estimating the intensity of convection. This is the basis of the moist convective adjustment scheme.

The moist convective adjustment (MCA) schemes (Manabe *et al.*, 1965; Miyakoda *et al.*, 1969; Krishnamurti and Moxim, 1971; Kurihara, 1973) are the simplest cumulus parameterizations. In MCA, it is assumed that deep moist convection acts to restore the lapse rate to a neutral or stable condition, and that there exists a critical temperature and moisture profile associated with the neutral or stable state. When the large-scale sounding becomes more unstable than this critical state, it is adjusted toward the critical state. This stabilization is assumed to be caused by cumulus convection.

The MCA schemes can be separated into hard and soft varieties. In the hard convective adjustment (according to Krishnamurti *et al.*, 1980), if some portion of a given column is convectively unstable, that is $\partial\theta_e/\partial z < 0$ for that region (where θ_e is the equivalent potential temperature), only this portion of the sounding needs to be adjusted to eliminate the instability. The value of the moist static energy ($h = gz + c_pT + Lq$) of the adjusted portion of the sounding must be the average moist static energy of the initial sounding over the unstable layer, in order to ensure energy conservation during the adjustment. Hard convective adjustment produces unrealistic modifications of the large-scale sounding by excessively cooling and drying the lower troposphere, and producing too much precipitation.

Because of the problems associated with the hard convective adjustment schemes, efforts have been made to improve them, e. g., by producing much slower and more realistic adjustment. Such methods are known as soft adjustment schemes. One of the soft schemes assumptions, according to Krishnamurti *et al.*(1980), is that the hard adjustment occurs over a fraction σ of the grid-scale area. Over the remaining area $(1-\sigma)$ it is assumed that the vertical profiles of temperature and humidity remain invariant during the time step. The final sounding in the σ region is determined from the construction of a moist adiabat. Then the final temperature and mixing ratio on the grid scale are just the average of those in the two regions, σ and $(1-\sigma)$. Although this represents an improvement over the hard convective adjustment, because of the occurrence of the maximum instability before the time of maximum convection in the tropics, the soft adjustment scheme shows a lag of 1 to 2 days between the calculated and observed precipitation. This large lag makes the soft adjustment scheme a poor choice when the timing of precipitation is important.

Kuo (1965, 1974) designed a cumulus parameterization based on the relationship between convective rainfall and large-scale moisture convergence. Since observational studies have shown that there is a strong correlation between the observed convective rainfall and the total large-scale convergence of water vapor in the column, Kuo chose the large-scale moisture convergence rate as a key variable to parameterize the effects of convection in large-scale models. Parameterizations (Kuo, 1965, 1974; Anthes 1977; Krishnamurti *et al.*, 1976, 1980, 1983; and Molinari, 1982) based on large-scale moisture convergence are called Kuo schemes. Basically, Kuo (1974) assumed that a fraction, $(1\text{-}b)$, of the total water vapor converge is condensed and precipitated, and so heats the column, while the remaining fraction, b, is stored and acts to increase the humidity of the column. Determination of b is obviously an important aspect of the Kuo scheme. Several methods have been proposed to determine b, by different authors. The vertical profiles of convective heating and moistening are based on the assumption that environment is modified through the mixing of cloudy air and the environmental air. The convective condensation heating and moistening are, therefore, directly proportional to the local excess of cloud temperature and moisture over the corresponding environmental values. The cloud temperature and moisture content can be calculated from moist adiabatic processes, although they also can be calculated from a cloud model (Anthes, 1977). The results of the Kuo scheme are much improved over those of the moist adjustment schemes, and therefore, the Kuo scheme has been popular.

An advantage of Kuo's scheme is that it provides immediate measures of cumulus-scale heat and moisture fluxes in terms of the measurable large-scale variables, without having to compute cloud dynamical processes and cloud microphysical processes. On the other hand, however, the simplicity of the Kuo scheme makes it impossible to see explicitly the interactions between cumulus clouds and the large-scale motions. Also because of the simplicity, many factors have to be determined empirically. Whereas convection is strongly controlled by the large-scale convergent flow in the tropics, it seems to have less significance in the extratropics (Frank, 1983; Tiedtke, 1989). This makes applications of the Kuo scheme in the extratropics questionable. More importantly, objections have been raised against Kuo's assumption that the environment is heated by the mixing of cloud air with the environmental air. As has been shown by Yanai *et al.* (1973), Ogura and Cho (1973), cumulus convection interacts with environment mainly through cumulus-induced subsidence in the environment between clouds, rather than through the mixing of cloud air with the environmental air.

Arakawa and Schubert (1974) developed a sophisticated cumulus parameterization scheme which includes many physical processes. In the Arakawa-Schubert (AS) scheme, a spectrum of cloud types is considered, so that the effects of different cloud types can be seen explicitly. Also, the AS scheme relates convective clouds to the large-scale forcing, which involves horizontal and vertical advection, radiation, and the surface fluxes of heat and moisture, rather than only large-scale moisture convergence as in the Kuo scheme. In particular, the AS parameterization makes the use of the assumption that the real atmosphere is in a quasi-equilibrium state, in which the rate of destabilization by large-scale processes and the rate of stabilization by cumulus convection almost balance each other. That is, the large-scale forcing produces convective clouds, and the clouds consume the instability caused by the large-scale forcing, making the atmosphere stay close to an equilibrium state in which the instability of the atmosphere remains nearly unchanged. In this way, the AS parameterization is an adjustment scheme. The intensity of convection, expressed in terms of the cloud mass flux, is determined by the large-scale forcing. Then, through the spectral cloud model, other cloud properties can be determined. With these cloud properties known, the effects of cumulus convection on the environment can be determined. The

AS scheme assumes that an ensemble of cumulus clouds affects its environment in two major ways: (1) by inducing subsidence between clouds, which warms and dries the environment; and (2) through detrainment of the saturated air, which contains liquid water or ice, from cloud top. Evaporation of the detained cloud water causes cooling and moistening of the environment.

Since the AS scheme relates cloud activity to the total large-scale forcing, not just the large-scale water vapor convergence as in the Kuo scheme, it is more realistic. The AS scheme is also appealing because of its clear physical concept of the interaction of cumulus clouds and the large-scale environment. In this respect it is presently not matched by any other scheme.

An obvious disadvantage of the AS scheme, however, is its complexity, which makes it more difficult to implement into large-scale models, and computationally more expensive. Although the key assumption of AS scheme, the quasi-equilibrium hypothesis, has received considerable support from observations (e.g., Lord and Arakawa, 1980; Lord 1982; Arakawa and Chen, 1987; Xu and Emanual, 1989) and numerical simulations (e.g., Ogura and Kao, 1987; Kao and Ogura, 1987; Grell *et al.*, 1991; Xu and Arakawa, 1992), there are still some questions about the validity of the quasi-equilibrium assumption for mesoscale and non-slowly varying fields (Frank, 1983; Tiedtke, 1989). Besides, the AS scheme assumes that the instability increase due to the large-scale processes will immediately be released by convection, so that the atmosphere will not increase its instability. This means that the AS scheme cannot predict the instability stored in the weather system. In the middle latitudes, however, sometimes the instability produced by large-scale processes can be stored in the atmosphere without triggering convection. Using the AS scheme in forecasting models for such cases will not produce accurate results.

Betts (1986), Betts and Miller (1986) presented another cumulus parameterization in which the main assumption is that when the sounding shows some kind of convective instability, it is adjusted by convection toward a quasi-equilibrium reference state. This scheme therefore belongs to the moist adjustment type schemes. However, in contrast to the previous adjustment schemes, the adjustment profiles have been chosen to represent the thermodynamic structures which are typically observed in convective situations and which resemble quasi-equilibrium states between the large-scale forcing and cumulus convection. Another important difference between the Betts-Miller scheme and the traditional adjustment schemes is that the adjustment is applied over a finite time interval, which makes the scheme a relaxation scheme. These two critical differences make the results from this scheme much better than those of the traditional adjustment schemes (Tiedtke, 1989). Also, the scheme is very simple, since all it needs are the specified reference profiles, the relaxation time, and a criterion for activating the scheme.

The key limitation of the Betts-Miller scheme is the definition of the reference profiles. Betts and Miller (1986) specified the reference profiles empirically from observed soundings. It is impossible that the reference profiles are unique; they must differ from region to region and may also depend on the synoptic situation. In addition, as mentioned before, the quasi-equilibrium assumption may be valid only for slowly varying fields but not for faster systems, making the scheme not applicable in meso-scale models. Also, as for most other adjustment schemes, the interactions between cumulus clouds and the large-scale processes are not explicitly shown.

Since penetrative downdrafts have been shown to make important contributions to the large-scale heat and moisture budgets (Johnson, 1976, 1980; Nitta, 1977, 1978), their effects were included in most of the cumulus parameterization schemes mentioned above (e. g., Cheng and Arakawa, 1990; Betts and Miller, 1993), as revisions of the original schemes.

We defined a Generalized Convective Available Potential Energy (GCAPE), and devised a parcel-moving algorithm for calculating it. The GCAPE of GATE data and ASTEX data were calculated, and the effects of ice were included. We found a high positive correlation between the rate of GCAPE production by large-scale processes and the observed precipitation rate, while a negative correlation exists between the GCAPE itself and the precipitation rate. The time change rate of observed GCAPE is much smaller than the GCAPE production rate by large-scale processes, implying that the real atmosphere stays close to a neutral state.

We also devised a penetrator algorithm, in which mass transport occurs in penetrative updrafts (downdrafts) with compensating layer-by-layer sinking (rising) motion. The solution was obtained using nonlinear optimization theory. Our results show that the penetrator algorithm is more effective than the Lorenz algorithm, in certain respects . It detects more GCAPE, and does not require high vertical resolution. The liquid water / ice distribution within reference-state cloud layers obtained with the penetrator algorithm is much smoother than that obtained with the Lorenz algorithm. Downward penetrators have been detected, but our results show that their contribution is much smaller than that of the upward penetrators.

The simplest cumulus parameterizations are the moist convective adjustment (MCA) schemes (Manabe *et al.*, 1965; Miyakoda *et al.*, 1969; Krishnamurti and Moxim, 1971; Kurihara, 1973). In MCA, it is assumed that deep moist convection acts to restore the lapse rate to a saturated moist adiabat, which can be called the "equilibrium state." When the large-scale sounding becomes more unstable than the equilibrium state, and if sufficient moisture is available, the sounding is adjusted toward the equilibrium state. This stabilization is attributed to cumulus convection. The main limitations of MCA are that it does not simulate penetrative convection, and that the equilibrium state is saturated and so not very realistic.

Arakawa and Schubert (1974) developed a sophisticated cumulus parameterization which includes many physical processes. It can be viewed as an adjustment scheme. In the Arakawa-Schubert (AS) parameterization, a spectrum of cloud types is considered, so that the effects of different cloud types can be seen explicitly. Also, the AS parameterization relates convective activity to the large-scale forcing, which involves horizontal and vertical advections, radiation, and the surface fluxes of sensible heat and moisture. In particular, the AS parameterization makes use of the assumption that the real atmosphere is in a quasi-equilibrium state, in which the rate of destabilization by large-scale processes and the rate of stabilization by cumulus convection almost balance each other. That is, the large-scale forcing produces convective clouds, and the clouds consume the instability generated by the large-scale forcing, so that the atmosphere stays close to an equilibrium state in which the conditional instability is weak, or non-existent. In this sense, the AS parameterization is an adjustment scheme.

According to the quasi-equilibrium hypothesis, the rate of instability increase due to large-scale processes is fully and immediately counteracted by convection, so that the atmosphere does not become very unstable. The assumption of such a quasi-equilibrium means that the AS parameterization cannot predict the Convective Available Potential Energy (CAPE) stored in a weather system. Some "relaxed" schemes, in which the exact quasi-equilibrium assumption is not strictly enforced, have been developed to implement the AS parameterization (e.g. Moorthi and Suarez, 1992; Randall and Pan, 1993). These schemes adjust towards the equilibrium state over a finite time scale.

Betts (1986), Betts and Miller (1986) presented a "relaxed" convective adjustment scheme in which, as in the other adjustment schemes, a conditionally unstable sounding is adjusted by convection toward an equilibrium state. They specified the equilibrium temperature sounding to follow a virtual moist adiabat at low levels and a pseudoadiabat at high levels. They specified the equilibrium moisture profile empirically, although in fact it may vary for different regions and synoptic situations. The feedbacks between cumulus clouds on the large-scale environment were not explicitly or "mechanistically" represented, e.g. in terms of mass fluxes.

Since the fundamental physical basis of adjustment methods is that convection acts to release the convective instability (or conditional instability) so as to drive the atmosphere towards a neutral state, a measure of the conditional instability is a key ingredient of such schemes. The conventional methods of measuring conditional instability are not fully satisfactory, however: the effects of environmental return flow are neglected, and the level of origination of the lifted parcel, must be assumed. The Generalized Convective Available Potential Energy (GCAPE) of Randall and Wang (1992) overcomes these restrictions, and therefore is a prior more accurate measure of the conditional instability. Based on Lorenz's (1978, 1979) concept of Moist Available Energy (MAE, Randall and Wang 1992) defined the GCAPE as the "vertical component" of the MAE and used it as a measure of the conditional instability of an atmospheric column. The definition of the GCAPE makes a "reference state," which is the unique state in which the system's enthalpy is minimized; it is also a statically neutral or stable state.The GCAPE is the vertically integrated enthalpy difference between a given state and the corresponding reference state, and represents the total potential energy available for convection in a given sounding.

In this paper, we propose an adjustment scheme based on the concept of GCAPE. The new parameterization combines some elements of the Arakawa-Schubert parameterization and the Betts-Miller parameterization, and tries to correct some limitations of those two parameterizations. The reference state associated with the GCAPE is chosen as the end-state of the adjustment, or the equilibrium state. This equilibrium state is determined by the given state, so that it varies in space and time. We relax towards the equilibrium state (as in the Betts-Miller parameterization), so that no strict quasi-equilibrium between large-scale forcing and convection is imposed.

We are attracted to the idea of using the GCAPE reference state as the equilibrium state of the adjustment because the GCAPE reference state is completely general and is not based on a cloud model. We have to keep in mind, however, that the GCAPE reference state is reached by reversible adiabatic processes. Real convection involves crucially important irreversible processes such as precipitation and mixing. Obviously, a cumulus parameterization has to take these irreversible processes into account.

We take them into account by using a simple cloud model. This means that although we avoid the use of a cloud model in the definition of the equilibrium state, we do use one to determine the convective feedback.

One might argue that an ideal cumulus parameterization would avoid using any cloud model at all. This is the idea behind the Betts-Miller parameterization. It is also the idea behind the empirical cumulus parameterization developed by Liu (1995), who used the logical framework of the AS parameterization, but employed both an empirical equilibrium state (in which an empirically defined measure of CAPE is small) and an empirical formulation for the feedback of the convection on the large-scale fields.

It seems desirable to avoid both empiricism and cloud models as far as possible. The present study aims to show that it is possible to use the concept of GCAPE to define the equilibrium state without using empiricism or cloud models, but we do resort to a cloud model to determine the convective feedback.

There is no contradiction between the use of the idealized reference state which is defined with respect to adiabatic reversible processes and the simultaneous use of a cloud model which includes irreversible processes like mixing and precipitation. Our idea is that the convection "tries" to adjust to the reference state, but that irreversible processes prevent this adjustment from being fully realized.

The incorporation of a cloud model inevitably and regrettably causes our parameterization to fall far short of the power and generality of Lorenz's MAE concept. For example, the cloud model does assume particular levels of origin for the updrafts and downdrafts.

As explained in detail later, we use the predicted (or observed) sounding and the corresponding GCAPE reference state, together with a relaxation time scale (discussed below), to determine the convective tendency of the moist static energy. This is not enough for a cumulus parameterization, however. In a prognostic model we need to know the tendencies of temperature and moisture separately.

In order to find them, we introduce the cloud model mentioned above, the form of the diagnostic model of Nitta (1975), modified to incorporate the downdrafts of Johnson (1976).[1] The convective moist static energy tendency is used as input to the diagnostic model, which

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[1] The cloud model used by Nitta is an entraining plume model of the type used in the AS parameterization. Emanuel (1991) has criticized the use of entraining plume models in cumulus parameterizations. Recent work by Lin (1994) suggests, however, that these models can in fact serve as realistic agents of convective transports.

determines the corresponding tendencies of temperature and moisture, and also yields the precipitation rate.

A second new aspect of our parameterization is that we relate the adjustment time scale to the large-scale forcing, so that the intensity of cumulus convection is controlled by the large-scale forcing (as in the Arakawa-Schubert parameterization).

A cumulus parameterization for large-scale models has been presented. It is an adjustment scheme. The reference state associated with the GCAPE is the end-state of the convective adjustment. This reference state varies from case to case, depending on the given soundings. The time scale for the adjustment also varies, ranging from several hours to several tens of hours, depending on the intensity of the large-scale forcing. Because the adjustment time scale is related to the large-scale forcing, the intensity of convective activity is determined by the large-scale forcing as in the Arakawa-Schubert parameterization.The methods of Nitta (1975) and Johnson (1976) are combined to diagnose the convective heating and drying rates.

The closure assumption of the present parameterization can be written as

$$\left(\frac{\partial h}{\partial t}\right)_{CU} = \frac{h_r - h}{\tau_{adj}}. \qquad (2)$$

Although (1) looks similar to the closure assumptions of Betts (1986) which are

$$\left(\frac{\partial T}{\partial t}\right)_{CU} = \frac{T_{ref} - T}{\tau_{adj}} \qquad (3)$$

and

$$\left(\frac{\partial q}{\partial t}\right)_{CU} = \frac{q_{ref} - q}{\tau_{adj}}, \qquad (4)$$

where T is temperature q is total water mixing ratio, and the subscript "ref" denotes the quasi-equilibrium reference state profiles, some important differences exist. One is that we allow τ_{adj} to change from case to case, depending on the large-scale forcing, while Betts and Miller (1986) use a prescribed constant τ_{adj}. In our parameterization, when the large-scale forcing is strong, τ_{adj} is small; and so the effects of convection are strong. On the other hand, when the large-scale forcing is weak or negative, a large τ_{adj} is used, and so convection is inhibited. In this way, the intensities of cloud activity and precipitation are related to the large-scale forcing.

The criterion for activating the Betts-Miller parameterization is that positive buoyancy is encountered when a hypothetical cloud parcel is lifted adiabatically from the boundary layer. However, as it has been shown (Thompson et al., 1979; Wang and Randall, 1994) that in GATE the observed precipitation rate is positively correlated with the intensity of large-scale forcing, but negatively correlated with the CAPE. This means that it may be more realistic to relate the effects of convection to the large-scale forcing than to the amount of CAPE.

Although both the Betts-Miller parameterization and the present parameterization are relaxation schemes, the final reference states are different. Betts (1986) determined the equilibrium state empirically from observed soundings. The equilibrium state of our parameterization is determined by the given soundings and the GCAPE theory, modified to include the effects of detrainment below the neutral buoyancy level.

A key difference between our parameterization and the Arakawa-Schubert (AS) parameterization is in the calculation of the cloud-base mass flux. In the AS parameterization, the quasi-equilibrium assumption is used to calculated the cloud-base mass flux. The quasi-equilibrium assumption requires that, at any moment, the rate of production of CAPE by large-scale forcing is balanced by the consumption of CAPE by convection, so that after each time step the CAPE remains unchanged. A cloud model is used to measure conditional instability and to define the reference state.

In our parameterization, no cloud model is needed to find the reference state. The effects of convection on the moist static energy are obtained from (1). Then, by using Nitta's method, we determine the effects of convection on the temperature and moisture fields. Both the AS parameterization and our parameterization relate the intensity of convection to the large-scale forcing, but in different ways. The present parameterization is a relaxation scheme in which no exact balance is required.

We do not adjust directly to the temperature, moisture, and condensed water of the reference state because this state is highly unrealistic, especially in view of high condensed water contents in the upper troposphere. We have considered the following strategy, however: On a given time step, adjust the temperature, moisture, and condensed water some fraction of the way to the reference state. Then, within the same time step, allow a microphysics parameterization to reduce the condensed water concentration in the upper troposphere by precipitation, and to increase the water vapor content of the lower troposphere by evaporating the falling rain. This approach would still include a "cloud model" in the sense that we would have parameterizations of precipitation, evaporation, and so on. Future work may go in this direction.

We regard this as an exploratory study. Certainly much additional work is needed before the ideas here are ready for application in large-scale models. Nevertheless, we are encouraged by our results to date and feel that this approach merits further investigation.

Additional information about our research and the results we have obtained is given in the attachments.

II. Human resources

This project supported a graduate student, J. Wang, who successfully defended his Ph.D. thesis in early 1994. He was then further supported as a Postdoctoral Researcher for about one year, until he found a job at the Scripps Institution of Oceanography.

III. Publications resulting from this project

Wang, J., and D. A. Randall, 1991: The Moist Available Energy of a Conditionally Unstable Atmosphere. Paper presented at the *19th Conference on Hurricanes and Tropical Meteorology of the American Meteorological Society*, Miami, Florida.

Randall, D. A., and J. Wang, 1991: The moist available energy of a conditionally unstable atmosphere. *Journal of the Atmospheric Sciences*, **49**, 240-255.

Wang, J., 1994: *Generalized Convective Available Potential Energy and Its Application to Cumulus Parameterization*. Ph.D. dissertation, Colorado State University.

Wang, J., and D. A. Randall, 1994: The moist available energy of a conditionally unstable atmosphere, II: Further analysis of the GATE data. *Journal of the Atmospheric Sciences*, **51**, 703-710.

Wang, J., and D. A. Randall, 1994: A cumulus parameterization based on the concept of GCAPE. Paper presented at the *Tenth Conference on Numerical Weather Prediction of the American Meteorological Society*, Portland Oregon.

Wang, J., and D. A. Randall, 1996: A cumulus parameterization based on the generalized convective available potential energy. *Journal of the Atmospheric Sciences*, **53**, 716-727.

Attachments:

Randall, D. A., and J. Wang, 1991: The moist available energy of a conditionally unstable atmosphere. *Journal of the Atmospheric Sciences*, **49**, 240-255.

Wang, J., 1994: *Generalized Convective Available Potential Energy and Its Application to Cumulus Parameterization*. Ph.D. dissertation, Colorado State University (cover page and abstract only).

Wang, J., and D. A. Randall, 1994: The moist available energy of a conditionally unstable atmosphere, II: Further analysis of the GATE data. *Journal of the Atmospheric Sciences*, **51**, 703-710.

Wang, J., and D. A. Randall, 1996: A cumulus parameterization based on the generalized convective available potential energy. *Journal of the Atmospheric Sciences*, **53**, 716-727.